Chapter 3: Petrologic evolution of Martian volcanism and clues from meteorites
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3.1 Introduction

The formation and evolution of the Martian surface, and ultimately the fate of its various volcanic regions, are revealed in the chemistry of Martian meteorites, and the mineralogy of the crust. The interpretation of the chemistry from remote sensing instrumentation from orbiters and landers/rovers has been strongly influenced by our knowledge of geochemical processes on Earth, the Moon, and other asteroidal, rocky bodies [1]. The composition, and thus evolution, of the Martian crust has been resultant from SNC (shergottite – nakhlite – chassignite) meteorites, which are the igneous counterpart of rocks of Martian origin [2, 3, 4]. These meteorites are chemically similar to terrestrial basalts and cumulates, with the exception of higher concentrations of iron and other volatile elements (i.e., phosphorous, chlorine, chalcophile) [5]. From Tuff et al. [2], most Martian meteorites have relatively young crystallization ages ranging from 1.4 billion years to 180 million years ago [6]. These are considered to be derived from young, usually lightly-cratered volcanic regions, such as the Tharsis volcanic region [6, 7]. Though mainly basaltic in composition, surface rocks are richer in nickel and sulfur with lower manganese/iron ratios than Martian meteorites [2], which can be explained by the differences in oxygen fugacity during melting of the sulfur-rich mantle [2, 8]. From this difference, Martian meteorites are tied to surface rocks through an early (3.7 billion years ago) oxidation of the uppermost mantle, producing the more recent volcanic rocks [2]. The variety of compositions from basaltic, olivine volcanic rocks through pyroxene-cumulate nakhlites and peridotite chassignites, enable the hypothetical reconstructions of the silicic evolution of Mars similar to techniques used on Earth [2, 5, 9].

In this chapter, we will explore the basic Martian composition of the crust and mantle in regards to volcanic regions, the crustal components of the North-South dichotomy, and more detail of the SNC meteorites in relation to the evolution of the mantle for volcanism.

3.2 Temporal evolution of volcanism

Mars is not uniformly distributed as observed by a diversity of volcanic landforms such as shield volcanoes, tholi, paterae, and small domes, as well as vast volcanic plains with various textures and lava flow patterns, as noted throughout this book. Such a diversity of volcanism comes from different eruption styles and changes of eruption through time (see also Chapter 4), though it is the localization of the various compositions and subsurface material (i.e., more silicic components could influence the overall ascension of mantle material to form different volcanic constructs and eruption style). The two topographically-dominating volcanic provinces are the Tharsis and Elysium regions, whose morphologies are analogous to basaltic landforms found on Earth [10].
Both are situated close to the dichotomy boundary between the cratere (older) highlands and the northern lowlands (see Section 2.2.2 of this Chapter). The Tharsis region consisting of Olympus Mons, Ascaeus Mons, Pavonis Mons, and Arsia Mons share many characteristics with Hawaiian basaltic shield volcanoes [11], as with their multiple lobate flows of lava [12, 13] and complex nested summit calderas (Chapter 4). Several authors [10, 14, 15] summarized the chronology and classification of Martian volcanic activity, grouping volcanic landforms into: (i) shield volcanoes; (ii) domes; (iii) highland paterae; (iv) volcano-tectonic features.

The temporal connection of extensive fluvial activity (e.g., valley networks) and volcanic episodes (highland paterae) is hypothesized from phreatomagmatic interactions (e.g., Tyrrhena Patera or western flanks of Elysium Mons) [10, 16]. Werner and Greeley and Spudis [10, 17] make note that the Hellas basin formation and the accumulation of highland paterae (at least of later-stage patera activities) were not triggered by the impact event itself. Back at the Tharsis region, there are notable differences in the structural content between Olympus Mons and Alba Patera, suggesting an environmental change [10]. Poulet et al. [18] deduced the geologic record of early Mars from a variety of features by: (i) widespread silicate-dominant formation and related alterations (e.g., phyllosilicates); (ii) existence of an active magnetic dynamo; (iii) erosion of extensive valley networks; (iv) large-basin scale impact cratering; and (v) construction of the Tharsis volcanic province. While an early Mars climate would have influenced these factors and subsequent features, volcanism is known to be a major factor in the long-term climate evolution of Mars, liberating volatile species from the interior to the atmosphere [19, 20, 21].

The massiveness of the Tharsis region has been implicated in a transition from a silicate-forming era (phyllosian) to a sulfate-forming era (theikian) [18, 22, 23]. Poulet et al. [18] stated that the timing of Tharsis volcanism is critical to understanding whether this change in mineralogical environments is plausible. While the bulk of the Tharsis region was probably emplaced in the Noachian [20], the load and emplacement on the regional lithosphere may have influenced the orientation of Late Noachian/Early Hesperian valley networks. This is further implied by evidence from crater counts and tectonic mapping from remote sensing techniques [18, 24, 25]. However, the notion of the Tharsis region being of Noachian-age has been disputed by [21] by the lack of craters on much of Tharsis, implying that this region to be Hesperian or Amazonian and requiring significant post-Noachian resurfacing. However, Poulet et al. [18] express a plausibility for these two contrasting views to be viable. From this, the majority of the Tharsis crust is constructed in the Noachian [26], but where extensive volcanic resurfacing persists through the Hesperian/Amazonian eras [27]. Parts of Tharsis from magnetization observations may also pre-date Hellas [18, 28]. The interpretation of the Tharsis construction near the end of the Noachian led to the silicate to sulfate transition may not be consistent with the bulk “old-age” of Tharsis [22, 23]. In the Hesperian, volcanic plain emplacement (particularly in the northern lowlands) resurfaced ~30% of Mars [29]. From Viking mapping, more than half of the volcanic resurfacing on Mars is early Hesperian or younger [30, 31].

3.2.1 Primordial Crust
Such volcanic transitions imply the changing of the Martian crust altogether, suggesting a need to differentiate the modern crust and primordial crust. The primordial crust would then be the crystallization of a magma ocean within the first 100 My after the formation of the Solar System [32]. In terms of volcanism, the first crust formed by early phases of intense partial melting of the mantle [33, 34], probably having a thickness of ~20 – 30 km [35]. However, the timing of the primordial crust is still under observation.

Isotopic anomalies preserved in SNC meteorites provide vital constraints about the timing of this pre-Noachian global scale differentiation within the deep Martian interior [36]. For example, the Hf-W-Th indicates that core formation occurred in the first 10 – 15 Myr [37, 38], though impacting to create large basins or even the dichotomy may accelerate this process (see Chapter 1 and 2). U-Th-Pb and Sm-Nd measurements show that the initial differentiation between the Martian mantle and primordial crust occur in the first 30 – 50 My [32, 39]. This supports the hypothesis where fractionation producing the core and the crust occurred sometime within the first 30 Myr [38]. The Sm-Nd petrology of shergottites provide a clear separation of at least two distinct Martian mantle reservoirs (depleted versus enriched in Rare Earth Elements – REE) in the first 10 – 100 My of Martian geologic history [32, 40]. An early magmatic build-up from the Martian highland crust required a rapid release of volatiles from the interior, forming an early atmosphere of Mars [41, 42].

The formation processes of the primordial crust remain debatable. The first hypothesis is the crystallization of a magma ocean (like that of the Moon), which would have been formed a resultant anorthositic crust [43]. However, this case seems unlikely given the larger size of Mars and its volatile content and anorthositic detections by CRISM (Compact Reconnaissance Imaging Spectrometer for Mars) would rely on a different mineralogical process [44]. The second hypothesis is the crystallization of a magma ocean as that for an early Earth [5], where plagioclase crystallization is delayed and solidification proceeds with denser Fe-rich cumulate overlying Mg-rich material, which creates a gravitational instability and mantle overturn [45, 46]. Thus, a basaltic crust would form by partial melting from adiabatic decompression of the rising Mg-rich cumulate. The third hypothesis is a partial melting of a fertile primitive mantle composition. This comes from the REE analysis of the NWA 7533 meteorite breccia, where low degree partial melting (~4%) of a primitive Martian mantle is sufficient to explain the formation of an enriched primordial crust and a magma ocean would no longer be necessary [47]. However, this would ultimately form a uniformly thick global layer [48].

The reshaping and ultimate burying (or eradication) of the primordial crust is also controversial. From [40], we know that there is no crustal recycling within the mantle as observed from the Sm-Nd signatures of shergottites. There are three potential processes that may have contributed to the processes of such reshaping of the primordial crust: (i) the formation of the North-South dichotomy dating back from the early Noachian, may have been synchronic with the primordial crust [42]. However, the nature of the dichotomy and its origin is still relatively unknown; (ii) the early Noachian period produced an intense volcanic activity time, leading to abundant crustal formation and partly burying the ancient primordial crust [33, 49]; and (iii) differentiation-fractional
crystallization occurred within a thick impact melt formed by a gigantic impact ~4.43 Ga ago, leading to alkali-rich clasts as observed in the meteorite breccia NWA 7533 [47].

3.2.2 Dichotomy

The North-South dichotomy is one of the most distinct textural differences observed on the surface of Mars. The southern hemisphere appears older due to the highly-cratered surface (~40%) and higher in elevation (by ~6 km) than the northern hemisphere (Figure 1). This dichotomy dates back between 4.5 and 4 Gyr [50], where age is determined from density crater measurements [51, 52, 53]. The bulk crust is basaltic, implying that a thick crust in the southern hemisphere should compensate for its relatively higher elevation [35]. However, there is a discrepancy regarding the density of such a thick crust compared to the geoid-topography ratio from [3-old] and may lead to lower crustal flow [54, 55]. A lighter crustal component comprised of feldspathic/anorthositic materials in relation to a thinner crust might be buried under subsequent basaltic volcanic products in the southern highlands [56], matching current geophysical data measured by orbital instruments. Crustal density values assumed from gravity and topography data are significantly lower (2,700 – 3,100 kg/m$^3$) than those of a purely homogeneous basaltic crust (3,100 – 3,700 k/m$^3$) inferred from meteorites and mineralogical observations from orbiters [56].

Figure 1: Latest chronostratigraphic map of Mars with noted surface ages. Names throughout this chapter are noted. Black dashed line indicates the major dichotomy. Blue circles are feldspar-rich zones [44, 57, 58]. Yellow circles indicate plagioclase-rich zones [59, 60]. Adapted from [61].
In geologic detail, the majority of the southern highlands is dominated by brecciated terrains that are mid-Noachian in age with some localized areas of early Noachian age. These zones of early-Noachian surfaces form isolated highland massifs, whereas the mid-Noachian regions are included in the Circum-Hellas Volcanic Province (CHVP), and the late Noachian terrains are found in Syrtis major [61]. Later Hesperian terrains (3.7 – 3.1 Ga) are found in the large volcanic provinces of Hesperia Planum.

The igneous rocks, as revealed by orbital spectroscopy techniques, are dominantly basaltic with a relatively high LCP/HCP (low Ca-pyroxene – high Ca-pyroxene) ratio and olivine phenocrysts [62, 63]. The scattered Noachian terrains are rather phyllosilicate-rich [64, 65]. Any remnants of an ancient crust should be located in the deepest places of Mars, or likely exposed at central uplifts of large impact craters and deep canyons, such as Valles Marineris [66, 67]. Resurfacing processes from constant volcanic activities likely complicate the identification of most ancient crust. Early Noachian terrains do show evidence of volcanic lava flows and pyroclastic deposits [68, 69, 70; see Chapter 1, 4 and 9].

In the Aeolis region along the dichotomy boundary at Gusev crater, where the Spirit rover landed in 2004, is the highly cratered Noachian southern highland and south of the Medusae fossae formation. The crater floor is covered by a Hesperian basaltic flow [71] containing olivine phenocrysts comparable to olivine-rich shergottites [72, 73]. This volcanic flow area encloses elevated terrains of the Columbia hills as the first alkali-basalt lithology identified on Mars [74]. However, the stratigraphic sequence is still unclear, so this could either represent a central uplift of older Noachian material or post-date the crater by infilling of Hesperian olivine-basalt lava flows [75]. Another rover at the dichotomy boundary, Curiosity at Gale crater in 2012, was the first mission to probe such ancient terrains (older than Late Noachian) [76]. From several analyses with the rover’s suite of spectroscopic instruments, mafic and light-toned felsic igneous rocks have been analyzed [57, 77, 78, 79, 80]. This included magmatic suites of: basalts, gabbros, norites, trachy-andesites, trachytes (alkaline-trend); and diorites and quartzo-diorites (sub-alkaline trends) [57, 77]. Such a diversity in the chemistry can lead to a variety of speculations in the magma processes, ascension dynamics, and overall volcanic structure (discussed in Chapter 1 and 4).

These observations raise several key questions, such as: Are these specific areas corresponding to localized plutons or an extensive ancient, differentiated crust? Are these felsic outcrops part of an excavated layer by huge impacts and younger basalt lava flows? Might such an evolved layer be formed by deep melting of a thickened basaltic crust preceded by crystallization of a magma ocean, or rather differentiated magma derived from low-degree partial melting of a primitive mantle?

3.3 Mineralogy in volcanic regions

The composition of magmas, produced by partial melting of planetary interiors, has the probable preservation of pressure-temperature conditions of melting throughout geologic time. For Mars, surface chemistry data must be obtained, such as thermal spectrometers or the gamma-ray
spectrometer onboard Mars Odyssey [81, 82], although this method is limited to more extensive regions that show clear morphological evidence of volcanism. Orbital spectroscopy has provided a complementary view on the mineralogy of magmatic rocks, showing that Martian igneous rocks have a mineralogy similar to that of Earth: dominated by plagioclase [83], pyroxenes [84, 85, 86], and olivine (sometimes lack thereof) [87]. The OMEGA (Observatoire pour la Minéralogie, l'Eau, les Glaces, et l'Activité) and CRISM imaging spectrometers provide evidence for a higher proportion of LCP in the Noachian highlands relative to the Hesperian-dated Syrtis Major [62, 84].

Baratoux et al. [33] stated there is a clear transition in the nature of pyroxenes observed in magmatic rocks and a possible increase in olivine at the surface of Mars over geologic time. This transition is most prominent near the Noachian/Hesperian boundary [84, 85]. LCP-rich outcrops have been suggested to be associated with higher degrees of partial melting, or trace the crystallization of the magma, followed by HCP-rich volcanism in the Hesperian [62]. Olivine-rich young volcanism and the evolution of the LCP + HCP ratios observed at the Hesperian/Noachian transition would be a natural sequence (and consequence) of primary melt compositions over geologic time, assuming that primitive magmas are a dominant component of the Martian crust [33, 72, 82]. Modeling of crystallization products of liquids produced by partial melting of a possible Martian mantle from conditions of the earliest Noachian era to the most recent Amazonian times have been attempted [33, 88, 89]. Such a transition is interpreted to be the consequence of the thermal evolution of the mantle. In that case, according to Baratoux et al. [33], Noachian rocks exposed at the surface are indeed linked to volcanism rather than being associated with mantle overturn following the crystallization of a magma ocean.

On Mars, the primary silicate paragenesis is of basaltic to andesitic-basaltic composition, as observed by several orbital observations such as Mars Global Surveyor (MGS), Mars Odyssey, and Mars Express [59, 90, 91, 92, 93], in addition to in-situ measurements by the Viking landers and rovers [94] as well as SNC meteorite analyses [95]. Main silicate types observed on the surface are olivine pyroxene and plagioclase, although silica-rich feldspathic minerals [96] indicate magmatic differentiation. Mars Global Surveyor (MGS) Thermal Emission Spectrometer (TES) has also detected plagioclase on Mars [92, 97]. Although silicate mineralogical data alone cannot support any one specific Martian magmatism model over another, data does support the basic igneous reservoirs proposed for Mars and used to constrain specific petrogenetic models [98].

Iron phases are also presumed on the surface of Mars [99]. The Martian regolith contains ~20 wt.% Fe3+ mostly under the form of iron oxides [100, 101]. Such iron phases have been confirmed by in-situ measurements from Viking 1 and 2 at Chryse Planitia and Utopia Planitia, Mars Pathfinder at Ares Vallis, and Spirit (at Gusev Crater) and Opportunity (Meridiani Planum) [92]. Columbia Hills, while having an age uncertainty compared to the 3.6 Gy-Hesperian olivine-basalt mineralogy of the Gusev crater floor [71, 75, 102], it has been noted that the tephrites are comparable in alkali constituents to some of the effusive alkali lavas (e.g., basalt, trachyandesite, trachyte – Figure 2) described in Gale crater [72, 75]. However, these alkali-rich rocks have not been observed (so far) in younger Amazonian shergottites, which are essentially olivine-basalts and detected in lava flows throughout the southern hemisphere [103].
Low-albedo (dark) regions on Mars provide the best opportunity for determining the composition of newer, pristine crustal lithologies [63]. Many of these regions are associated with volcanic provinces (e.g., Syrtis Major volcanic plateau) and analyses of their spectral properties from orbiters show a mafic/ultramafic composition [105, 106, 107, 108]. Such compositional analyses can reveal the eruption and emplacement of magma, even in understanding the crystallization processes that may occur [63, 109]. Besides being ultimately mafic and basaltic in nature, there are also altered materials on the surface of Mars. Interestingly, Mustard et al. [63] observed that relatively unaltered surface materials were found in volcanic terrains of Syrtis Major, Valles Marineris, Ophir Planum, and Sinus Meridiani. As stated early regarding the observation of two-pyroxene mineralogy consistent with SNC meteorites indicates that (i) the SNC mineralogies are representative of large volcanic regions of Mars as old as 2 – 3 Ga; (ii) the mantle was depleted in aluminum; and (iii) there has been little evolution in the composition of mantle sources [63].

LCP-rich outcrops have been identified in the Noachian walls of Valles Marineris [110], while higher ratios of HCP/LCP have been found in the Hesperian volcanic deposits of the Circum-Hellas Volcanic province [111] and Chryse and Acidalia Planitia of the northern plains [112]. Although dust is an extensive issue for properly characterizing the younger volcanic regions, there have been observations of how LCP concentrations at certain locations, such as the Amazonis Planitia [113], Noctis Labyrinthus, Echus Chasma [114], and volcanic domes in the northern plain
[115], all strongly dominated by HCP and plagioclase. Amazonis Planitia is noted to also be rich in olivine, as typified by the effusive volcanism in the area [113].

At Terra Tyrhenna, dated from early to mid-Noachian (see Chapter 9), CTX and HiRISE images observed bright rocks, where the CRISM instrument detected > 95% of feldspars in a nearby crater, which have been interpreted as anorthosites [116]. However, Rogers and Nekvasil [117] stated that sub-millimeter feldspars detected in the near-infrared spectra (such as CRISM) contain a mixture of at least < 50 vol% mafic minerals. The co-existence of Al-rich clays (i.e., kaolin) are likely formed by the alteration of highly felsic parent rocks [116], which have been detected at Xanthe Terra. Additionally, northeast Noachis Terra has been observed to have felsic polygonal light-toned outcrops. South of Xanthe Terra, light-toned massive bedrocks have been observed by HiRISE [60] and are dominated by LCP by CRISM data. Flahaut et al. [60] suggested that these LCP-rich rocks are potentially a Noachian pristine crust exposure. Fractional crystallization at low pressure of a basaltic melt can explain these felsic detections. CRISM data revealed that these terrains with feldspar-rich materials along with Al-silicates likely formed by the alteration of the parent felsic rocks [57]. Either way, the occurrence of feldspar-rich rocks points out early evolved magmatism on Mars. The occurrence of such felsic outcrops within large mafic terrains may be implied as buried ancient evolved crust or localized plutons.

Terra Sirenum and Terra Cimmeria, dating from early to late Noachian, display the strongest crustal remnant magnetism ranging from -100 to +100 nT/deg measured by Mars Global Surveyor [118]. A geomorphological study by Xiao et al. [49] has revealed these could be the early Noachian volcanic edifice remains are likely shield-like volcanoes with 50 – 100 km diameter and 2 – 3 km high. Terra Sirenum was also observed to have Fe/Mg phyllosilicates, hypothesized to be the alteration of an ancient Noachian crust rich in volcanic and impact-generated mafic glasses [119, 120], although CRISM detections of Al-silicates likely reflect the alteration of felsic parent rocks [121].

Few mineralogical data exist for plain-style volcanism with which to constrain their composition independently from the rheological evidence of broad lava flows [114, 122] suggesting a high fluidity (see Chapter 4). East of the large volcanoes of Tharsis is Echus Chasma, the source area of the Kasei Valles outflow channels [114]. The Echus-Kasei valley system is partially filled by lava flows originating from the main volcanic centers from the west [123], though no vent is observed [114], suggesting that localized volcanic fissures exist [124] or that Kasei Valles branched out from Valles Marineris [70 and references therein]. Echus Chasma and Eastern Noctis Chasma have platy-ridged textures, as observed by Mangold et al. [114]. Both areas display a mineralogy of basaltic lava flows with predominant plagioclases and pyroxenes. Models of platy-ridged lava flows generally show a viscosity in the range of 100 – 1000 Pa s [125]. Such low viscosities and yield strengths less than 200 Pa are usually found in the central Elysium Planitia [126], where extensive platy flows have also been reported [114, 125]. While Mangold et al. [114] argue that all platy-ridge flows may not be magmatic in origin, the textures at Echus and Noctis Chasma may indeed be associated with basaltic lava flows. Low viscosity and low yield strengths for silicate magma may be related to different parameters (see Chapter 4), including increased temperatures or low proportions of silicates.
3.4 Meteoritic evidence for volcanism

3.4.1 SNCs

Crater-forming impact events on Mars can generate sufficient energy to eject fragments of the crust through the atmosphere and into space (with an escape velocity ~5 km/s) [127] through near-surface bombardment [29]. Fragments as a result of these ejection events represent the currently recovered Martian meteorites that have been found on Earth in places such as Antarctica, Morocco, Libya, Egypt, Tunisia, France, Chile, USA, India, Nigeria, Mali, Mauritania, Oman, and Brazil [128].

Martian meteorites are divided into three main groups: shergottites (basalts of pyroxene and plagioclase, ~82% of known samples), nakhlites (Ca-rich clinopyroxenes with minor olivine, ~14%), and chassignites (greater amounts of olivine, ~3%), all known collectively as the Shergottite-Nakhlite-Chassignite (SNC) [33, 98, 129, 130]. Shergottites also have different subtypes, such as poikilitic, gabbroic, basaltic, and olivine-phyric. Additional rock types from Mars includes Lherzolites (coarse-grained ultramafic rocks of olivine and pyroxenes) and Orthopyroxenite (cumulate rock almost entirely of orthopyroxene) [98]. The readers are encouraged to refer to (https://imca.cc/mars/martian-meteorites-list.htm) for the most up-to-date list of Martian meteorite samples. These have been verified to be of Martian origin based on their distinct oxygen isotope signatures as identified by the techniques of Clayton and Mayeda [131].

While Martian meteorites offer an opportunity to study a number of surface and subsurface localities, cosmic ray exposures [6, 132] relate the ejection age of the rocks to be limited to a few locations on Mars. Nonetheless, they provide petrological, mineralogical, and geochemical data, which could otherwise be difficult to ascertain by landers and orbital spacecraft.

SNC meteorites being labeled as specimens from Mars was observed from their young crystallization ages [6] and fractionated rare-earth patterns [133, 134], both indicators that a sample came from a planet-sized rocky body, in addition to the presence of trapped Martian atmosphere in shock-produced glass and basaltic matrix [135, 136]. Most Martian meteorites are geologically young, with shergottites chiefly being mid- to late-Amazonian in age (< 716 Myr crystallization ages), whereas nakhlites and chassignites dated at ~1.3 Gyr [6, 114, 130, 137, 138, 139, 140, 141, 142, 143]. This is consistent with the HCP-rich and olivine mineral assemblages predicted to crystallize from primary melts [33]. This is also comparable to the youngest volcanic areas on Mars.

However, the extensive dust overlying young volcanic regions and basaltic shergottites do not hold a very clear link [90, 114]. Most basaltic shergottites are richer in LCP (ratios from 1:1 to 18:1) [144, 145, 146] when compared to the proportions derived from OMEGA data [114]. From [114] there are young (Amazonian-aged) volcanic areas that show LCP/HCP ratios from 1:4 to 1:6.
Considering the chemistry of these Martian meteorites, REE abundances and isotopic data suggest a rather heterogeneous source region for the shergottites [147, 148].

All and all, Martian meteorites (along with orbital and in-situ results) revealed the occurrence of felsic rocks dating from early Noachian [104, 149]. Such a discovery leads to some interesting questions regarding the nature of felsic igneous materials possibly remaining in the primary crust. A fundamental question that arises concerns the fate of the aluminum. As noted early, it is unlikely that Mars had an ancient anorthositic crust similar to the Moon. Treiman and Lindstrom [150] suggest that a basaltic crust 120–150 km thick could in fact account for the depletion of Al, which is consistent with the alteration products in Martian dust, but not of anorthosite [151]. Various felsic and alkaline rock compositions were analyzed at the surface by the Spirit and Curiosity rovers [72, 76, 77, 78, 152, 153] and fractional crystallization (or integration-assimilation) could be probable processes that formed these rocks [128, 153, 154].

Due to dating constraints, these meteorites do not directly reveal the composition of the nearly Noachian crust, but rather younger surfaces. Several authors [6, 47, 155] however note that two specimens are aged from Noachian that could potentially tell us more about the nature of a primitive Martian crust: ALH8400 (“Allan Hills” orthopyroxene sample) [156] and the regolith breccias of NWA7034 (with several paired stone such as 7533/7445/7475/7907 etc.) [155]. ALH8400 has been dated to about 4.5 to 4.071 Gyr by [6, 33, 130, 157] obtained by Lu-Hf and Pb isotope data. Because of its cumulative nature and homogenized mineralogy by diffusive processes, this meteorite serves as a complex sample for the reconstruction of magmatic processes on early Mars.

Sub-alkaline magmatic constituents analyzed within Martian breccia samples (e.g., NWA 7034) and in-situ measurements from Pathfinder and Curiosity reveal the putative andesitic-rich protolith [149]. The Curiosity rover in particular detected quartz-dioritic rocks [78] and likely rhyolitic protolith in Gale crater [158]. While the spatial and temporal relationships of the various rocks are continuously being researched, these findings within the Martian southern hemisphere are rather linked temporally, pointing out a common early crust. The combination of Pathfinder and MGS-TES data indicate the presence of andesites [94, 159], and with the zircon inclusion studies from the NWA 7034 meteorite sample [149], the 4.54 Gyr proto-crust could be andesitic. Recently, the InSight lander and its seismograph instrument at Elysium Planitia in the northern hemisphere also supports an andesitic crust [160]. The formation of the andesitic proto-crust either being primarily extracted from a primitive mantle or through magmatic processes affecting an even earlier primary crust remains unclear.

The second type of felsic lithology is alkaline. The Curiosity rover at Gale crater analyzed a variety of rocks ranging from basalt to trachyte, including anorthoclase and sanidine throughout the Bradbury and Kimberly formations [78, 80, 161, 162, 163]. These alkali igneous rocks echo within the NWA 7533 regolith breccia meteorite [47, 164]. While the alkalic rocks within Gale crater have likely been formed by fractional crystallization from a basaltic melt [165], the alkali lithologies in the meteorite sample likely formed 4.43 Gyr by differentiation and fractional crystallization within an impact melt sheet, likely reworking the andesitic proto-crust 4.54 Gyr described earlier [47, 149].
Interestingly, the chemistry of plagioclase is considered a great record-keeper of the chemical (and physical) processes of basaltic volcanism, especially utilizing different elements that segregate into olivine and pyroxene [98]. Papike et al. [98] noted that there is no plagioclase detection in meteorite samples of Yamato-980459 or MIL-00346.

NWA 7475 is considered to be the first catalogued indurated-polymict-breccia from Mars that could be representative of the anciently cratered southern hemisphere [166], providing constraints on the nature of the primitive crust and possibly early magmatism. From several studies on this sample [47, 155, 164, 166], this sample has a mixture of effusive and extrusive igneous polymineralic coarse-grained clasts including both felsic and mafic constituents. A unique feature of this meteorite is the presence of leucocratic felsic igneous clasts, in which this identification is the first Noachian felsic clasts that have been identified as trachy-andesites, trachytes, and monzonites [47, 155, 164, 166]. The mafic clasts have been identified as norites and basaltic andesites, mostly formed by pyroxenes (including orthopyroxenes and augites), plagioclases with andesines and oligoclases, and trace amounts of Fe-Ti–oxides [47, 155, 166]. Interestingly, the orthopyroxenes are similar to the cumulate nature analyzed in the ALH84001 sample [164]. Although basaltic clast compositions are typically characteristic of mafic rocks (with SiO$_2$ < 49.5 wt.%) [167, 168], the sample’s Na$_2$O contents are higher than the average crustal composition, while the MgO concentrations are lower. Moreover, the exsolution features in coarse crystals (whether in noritic and monzonitic fragments or in mineral clast forms) point to slower cooling, which in turn means deep-seated plutonic origins. This breccia is dated at 4.43 Gyr, which could be representative of a reworked Noachian surface regolith, possibly excavated from several deep impact melts [164]. This early Noachian Martian breccia provides two interesting insights to the early stages of crustal formation: (i) a buildup of an andesitic proto-crust no more than 20 Myr after Solar System formation, and (ii) several impacts on this primordial crust weathered 100 Myr later to form deep melts where differentiation occurred, which formed more alkali-rich rocks in the upper portion.

### 3.4.2 Mantle Reservoirs

Within the framework of a possible magmatic ocean on early Mars in that the volcanic products of mantle overturned, the igneous mineralogy of surface rocks does not require a peak in volcanism around the Noachian/Hesperian boundary [33]. Volcanic gas composition strongly depends on the atmospheric pressure [169], namely sulfur-dioxide rich volcanic gases being produced. A decrease of atmospheric pressure at the Noachian/Hesperian boundary would therefore solidify the hypothesis that Hesperian volcanic activity is responsible for the formation of massive sulfate deposits [33, 170, 171, 172]. However, it remains unclear if shergottites and Martian magmas are sulfide-saturated (or under-saturated) and therefore how much of this sulfur can contribute via volcanism [173]. Experimental results from Righter et al. [173] have found that primitive shergottite magmas are capable of dissolving ~4000 ppm S at sulfide saturation. With the melting
of the Martian mantle and consequent volcanism, this provides more than enough sulfur for the sulfate-rich soil and deposits observed on the surface of Mars.

Several authors [33, 49, 174] concluded that the Early Noachian was the prime time of widespread volcanism associated with planetary cooling and most rocks exposed at the surface should be related to this period of intense volcanism. Martian meteorites have provided a means to modeling the Martian mantle and basaltic magmatism. During the main shield period on Mars and a large eruption volume of shergottite lavas, load is emplaced (unevenly) on the underlying lithosphere, leading to flexure and consequently the development of a flexural bulge [128]. Flexural bulges and moats are observed on Earth, most prominently at the Hawaiian chain volcanoes, and based on gravity, on Mars in the Tharsis volcanic province [175, 176]. As with terrestrial rejuvenated lavas, a previously depleted mantle has to be chemically altered in order to induce partial melting through decompression during lithospheric flexure. On Mars, this is required for the source of nakhlites based on their Sr/Nd isotopic ratios.

Shergottites from enriched, intermediate, and depleted sources have been calculated to originate from mantle sources with a range of potential temperatures, such as ~1750°C compared to Noachian rocks at Gale crater (~1450°C), representative of products from a hot mantle plume [177, 178]. Such geochemical and isotopic characteristics suggest the occurrence of at least four silicate reservoirs that formed during the primordial differentiation of Mars [98, 143].

Wadhwa and Borg [143] described them as: (i) Depleted Mantle Reservoir 1 – source of light REE (LREE) depleted shergottites, such as Yamato-980459, QUE94201, Dar al Gani 476, Sayh al Uhaymir 005. This reservoir also has distinctive Hf/W isotopic ratios; (ii) Depleted Mantle Reservoir 2 – source of all clinopyroxenites (Nakhlites). While also LREE-depleted, this reservoir type has characteristic Nd and Hf/W isotopic signatures; (iii) Depleted Mantle Reservoir 3 – source of ancient orthopyroxenite crust, representative of the ALH-84001 meteorite. This mantle reservoir type has notable Sm/Nd isotopic signatures compared to the other types; and (iv) Enriched Mantle Reservoir – either an oxidized crustal or mantle component that contributed to the LREE enrichment of shergottites, such as Shergotty, Zagami, Los Angeles, and NWA 1110.

The identification of such distinct mantle reservoirs (depleted versus enriched) from which the shergottites were derived has given rise to two general models [98]: (i) assimilation of crustal material by mantled basaltic magmas (Figure 3A); and (ii) derivation of basaltic magmas from multiple reservoirs (Figure 3B, C).

Figure 3A proposes that the Enriched Reservoir is located within the Martian crust and Depleted Reservoir 1 is the shergottite mantle source [179, 180, 181, 182, 183]. Both of these reservoirs would have been produced during the initial stages of differentiation through crystallization of a magma ocean [98]. In this particular scenario, there are varying degrees of crustal adaptation by mantle-derived basalts that which produces the isotropic signatures observed in shergottites [138, 143, 183, 184, 185]. This particular scenario has the advantage of explaining temporal separation and long-term isolation of the two reservoirs (crust versus mantle emplacement). Furthermore, this model provides a means in which enriched and depleted components could mix during basaltic generation and consequent eruption [98].
For Figure 3B and Figure 3C, the Depleted Reservoir 1 and the Enriched Reservoir are located in the Martian mantle [138, 143, 185, 186, 187]. These multi-reservoir models must be isolated within the mantle for at least 4000 Myr and only mixed at the time of melting to produce relevant shergottite isotopic characteristics [98]. A fundamental aspect of these multi-reservoir models is that they must have been produced by the crystallization of a deep Martian magma ocean [188]. [186] explains a two-stage model of the depleted, intermediate, and enriched sources (Figure 3B). Heat-producing elements (e.g., K, Th, U, etc.) are concentrated in the late-stage crystallization of the putative magma ocean and can be mixed by convection processes [98]. Figure 3B illustrates where melts rise with no (or little) crystallization until they enter magma chambers in the Martian crust. Crystallization processes would then commence, where various mixtures can be derived by fractional crystallization [183, 186] before being transported to the Martian surface. Shearer et al. and Symes et al. [183, 186] proposed that this process links the olivine-basalts (e.g., Yamato-980459) to pyroxene-basalts (e.g., QUE-94201) within each shergottite end-member (depleted versus enriched).

Blinova and Herd [187] proposed a more complex, three-stage model to account for the REE systematics of shergottites (Figure 3C). In this scenario, the source of the depleted shergottites is a mixture of upper mantle cumulates (nakhlite parental magmas) at 1.3 Gyr and a deep mantle lithology that was located near the mantle-core boundary [98]. These two distinctive sources progressed in Nd isotopic systematics due to their different Nd/Sm ratios [98]. The deep mantle source was then transported to the shallower mantle by a plume and experienced melting over a variety of pressure regimes. Such a mixture of melts of the plume and mantle residue produces the REE patterns of the depleted shergottites. The enriched reservoir is rather a late-stage Martian magma ocean cumulation emplaced in the upper mantle, but the lower mantle plumes could also provide a heat source to initiate melting for the enriched (E.g., Zagami, Los Angeles) and intermediate Martian basalts (e.g., LEW-88516, NWA 4797).

Although these illustrative models do not fully explain the timing of reservoir isolation and mixing processes, each do impose their own requirements on the mantle processes [98, 186]. With the isolation of the reservoirs, the reservoirs must have thermal boundaries that allow for the transport of heat, though not the exchange of mass, while the mixing must initiate melting [98].
Figure 3: Models of Martian magmatism as hypothesized by shergottite geochemistry. (A) Crustal assimilation model [182]. (B) Multiple mantle source [186]. (C) Multiple mantle source [187]. See text for more details for each scenario. Adapted from Papike et al. [98].

3.5 Summary and Open Questions

The multiple observations of a variety of mineralogy from orbiters, landers, and SNC meteorites reveal a complex history of Martian crustal evolution. The detections of early Noachian-aged felsic outcrops throughout the southern hemisphere suggests that there are remnant exposures of an extensive ancient andesitic crust deeply impacted early in Martian geologic history. Concerning the paradox of the crustal average density, a fully-basaltic crust would be considered too high [35, 189], therefore Wieczorek and Zuber [35] proposed the occurrence of large crustal proportions composed of felsic materials. This is also evident in Martian meteorites.

One of the major knowledge gaps for Martian meteorites is the location of their excavated origin. Comparison of meteorite ages with crater chronology of the surface can show distinct “ancient” versus “young” [190], but a better understanding of the ejection physics of the meteorites (and consideration of shock effects and isotopic compositions) strengthens the conclusions of possible origins [29, 191, 192]. Hartmann and Neukum [52] noted that less than 10% of the Martian surface is younger than 1 Ga, including Tharsis, Amazonis Planitia, and Elysium.

Another prominent feature on Mars is the hemispheric dichotomy from early Noachian time, when the earliest crust of Mars being more likely basaltic than anorthositic. Long-lived mantle upwelling
related to high heat production produced a thickened felsic alkaline crust localized in the southern hemisphere. Consistent and widespread volcanism have likely curried the Martian felsic crust, only to be observed on localized scales from impact crater excavation.

The SNC meteorites provide the most detailed geochemical and mineralogical data for Mars [63, 129]. However, it should be noted that the implications for a global understanding of planetary evolution and widespread volcanism are very much limited by the inherent uncertainties in the exact locations from which the SNCs were derived, and whether the current SNC catalog provides a representative sampling of the mineralogic diversity of the Martian crust [63].

There are several remaining questions regarding Martian meteorites and the geological evolution of Mars, such as: How compositionally variable is the Martian interior and surface? How did the magma ocean crystallize? How have the diversity of mantle and crustal sources changed over time? How has magmatic behavior, from fractional crystallization to accumulation, evolved over geologic time? Is there a way to resolve the contrast between meteorite samples and remote-sensing data? Nonetheless, the diverse mineralogy that such Martian meteorites can provide show us just how complex volcanism can be dependent on the minerals present throughout Martian geologic history.

References


[129] McSween HY. (1994), What we have learned about Mars from SNC meteorites. Meteoritics, 29(6).


